Climate Controls on High Sea Surface Temperatures

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Submitted to *World Resource Review* February 9, 1996
Revised June 11, 1996

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Keywords: Sea Surface Temperature, Global Warming, Climate Dynamics, Cloud Feedback, Ocean-Atmosphere Interactions

Abstract

The ocean's immense thermal inertia imparts a significant mediating effect on the Earth's weather and climate. This suggests that our ability to predict surface temperature changes over land under a scenario of climate change will only come from an understanding of the climate controls on the surface temperatures over the oceans. In addition, research over the last decade or more has shown that the seasonal, interannual and longer-term natural climate fluctuations in the equatorial and mid-latitude regions are particularly sensitive to the sea surface temperature distribution in the warm tropical oceans. In response to these implications, a significant amount of research has been undertaken over the last five years to pinpoint the mechanisms that regulate and/or limit the sea surface temperature in the tropics. These efforts have migrated from simple energy budget models to considerations of the radiative-convective nature of the atmospheric column, and have since examined the influences from the large-scale atmospheric circulation and the feedbacks from ocean dynamics. This review summarizes this progression by first pedagogically describing some of the earliest studies, then highlighting several of the more directed investigations that have followed in recent years, and finally concluding with a discussion of a few of the most recent results.

1. Introduction

In one of the earliest assessments of the stability of the Earth's climate, Budyko (1969) showed how small variations in external forcing could lead to extreme, and in some cases, catastrophic climate events. In Budyko's case, the variations in external forcing were due to small changes (~1%) in solar radiation. In more recent years, the most consequential external climate forcing being studied is the observed anthropogenic increases in carbon dioxide and other greenhouse gases in the Earth's atmosphere. The suggestion that these greenhouse gas increases may result in a detrimental climate change occurring over the next century has elicited significant public concern as well as considerable scientific interest (see review by Mitchell, 1989). One of the more practical questions associated with this potential climate shift concerns the magnitude and the manner surface temperatures will change. While the societal impacts resulting from these temperature changes stem most directly from those that occur over land, it is at least as important to concern ourselves with the details of surface temperature changes implied/predicted for the ocean due to the considerable moderating effect the ocean has on the Earth's weather and climate. This necessity begs the question then, what are the mechanisms that regulate and place upper limits on ocean surface temperatures?

In addition to the climate change scenario alluded to above, studies addressing the predictability of the Earth's natural climate variations, such as those associated with the El Niño Southern Oscillation (ENSO) phenomena (see Philander, 1991), have shown that the climates in the tropical and mid-latitude regions are highly sensitive to the sea surface temperature (SST) distribution in the tropics (see Glantz et al., 1991). This implies that our ability to predict climate on seasonal and longer time scales over a large portion of the Earth depends crucially on our understanding of the mechanisms that control SST, and in this latter case, which mechanisms dominate the process in the warm regions of the tropics.

Figure 1 shows the seasonal climatology of SST for the global tropics (Reynolds and Smith, 1994). Evident in the figure are the fairly strong SST gradients near the subtropical regions leading to the relatively high SSTs observed over a large portion of the deep tropics, including the western half of the Pacific and Atlantic Oceans as well as most of the Indian Ocean. Figure 2 shows tropical SST population distributions for various time scales, ranging from instantaneous samples to the long-term mean. These two figures illustrate that in the present-day climate, climatological SSTs rarely exceed 31°C. In fact, historical records indicate that over the last 100 million years, SSTs have rarely exceeded this value (Crowley, 1993). This stability, along with the discussion above, has motivated many climate researchers to examine the mechanisms that regulate tropical SSTs and account for this upper limit. What follows is a review of these efforts, beginning with a fairly didactic account of the earlier studies, following with a discussion of several of the more directed studies from recent years, and concluding with a brief summary.

2. Zeroth-Order Surface Heat Budget Analysis

The mechanisms that limit SST growth on the Earth were first directly addressed by Newell (1979) in a discussion of the relationship between global climate and the world's oceans. Newell's approach was to balance a model tropical sea surface heat budget using the following bulk aerodynamic formula to represent the behavior of the surface heat loss terms as functions of SST:

$$E \propto U\,[Q_s(T_s) - r\,Q_s(T_a)] \quad (1)$$

$$S \propto U\,[T_s - T_a] \quad (2)$$

$$L \propto \sigma\,T_s^4\,[0.56 - 0.095\,(r\,Q_s(T_a))^{0.5}] \quad (3)$$

In the above formulations, E, S and L represent the surface latent, sensible and net longwave heat fluxes, respectively. The first represents the heat removed from the ocean surface when water evaporates into the overlying atmosphere. The second represents the transfer of heat by molecular/turbulent conduction. The third represents the difference between the upward (nearly blackbody) emission from the sea surface and the downward emission of thermal radiation from the atmosphere. The latter is distinguished from the incoming solar radiation at the surface, the insolation, which is often referred to as the shortwave heat flux. T_s and T_a represent the sea surface temperature (SST) and near surface air temperature, respectively. Q_s represents the saturation specific humidity at the given temperature, σ is the Stefan-Boltzman constant, and r and U represent the near-surface relative humidity and wind speed values, respectively. To close the system, Newell further assumed no ocean heat transport and a model tropical atmosphere with clear skies, and fixed values for air temperature (T_a = 27^{o}C), relative humidity (r = 70%) and wind speed (U = 3 m s^{-1}). The resulting heat budget is reproduced in Figure 3. Newell's clear-sky insolation value and the sum of the heat loss terms (E+S+L) balance at about 31^{o}C. The balance is primarily between incoming solar radiation and outgoing evaporative heat loss, with evaporation increasing rapidly with increasing SST. This prompted Newell to conclude that evaporation was the primary mechanism that limited climatological SSTs to about 30^{o}C.

There are three features in Newell's analysis that are worth pointing out in more detail. First, all three heat loss terms increase with increasing SST, and thus they each represent a negative feedback on increased SST warming (e.g., $\partial S/\partial T_s > 0$). For example, as T_s increases, H increases linearly, which in turn tends to limit or remove the SST increase. The latter result is partly attributed to Newell's choice of bulk aerodynamic formula and partly due to his constraint of using constant near-surface temperature and relative humidity values (cf. Seager et al., 1995). Second, the heat input term (i.e. the solar insolation) is independent of sea surface temperature, and thus, does not exhibit any kind of feedback on SST. Third, Newell's conclusion is a statement concerning the maximum observed SSTs, i.e., the upper limits of the climatological distributions shown in Figure 2.

3. One-Dimensional Radiative-Convective Considerations

A modification to Newell's simplified model was made by Graham and Barnett (1987; hereafter GB) based on their analysis of the relationship between large-scale tropical deep convection and SST. They found a fairly rapid increase in the intensity/frequency of deep convection with SST in the range of about 26-28^{o}C that is fairly widespread over the global tropical oceans. The essence of this result is shown in Figure 4 using 14 years of Highly Reflective Cloud data from the global tropics (Waliser et al., 1993; see GB's Figure 5). The horizontal axis gives the SST value, the left vertical axis represents the frequency of occurrence of large-scale convective rainfall systems associated with each particular SST, and the right vertical axis shows the corresponding monthly population distribution of SST. The decrease in convection at the very high SSTs evident in the figure was not resolved by GB's data set which had significantly fewer data points than the data set shown here. Note that in the absence of this feature (discussed further in Section 4), the relationship suggests that higher SST implies greater hydrostatic instability, and thus, more deep convection. The ubiquitous cloudiness associated with these convective systems prompted GB to modify the simplified model of Newell to account for the cloudiness bound to be present at such warm SSTs. They assumed that the effect of these cloud systems would result in a mean albedo of about 30%[1], and thus, reduce the incoming solar radiation in Newell's model to a degree that the surface energy budget balanced at about 28^{o}C (See Figure 3). This temperature coincides closely with the mode in the observed population distribution of SST shown in Figures 2 and 4. From these results, GB concluded that cloud cover in addition to evaporative cooling placed "an upper limit on tropical SSTs." However, because these results apply to a selection process of the mode (i.e. the most typical rather than the maximum SST), it might be more suitable to refer to this as a statement concerning the mechanisms that regulate tropical SSTs.

An additional feature that can be introduced using the data in Figure 3 is the concept of "cloud radiative forcing"(CRF). In Newell's clear-sky model, the insolation at the surface is 341 Wm^{-2}, while the effects of clouds in GB's modification reduces this value to 239 Wm^{-2}. The net effect on the shortwave heat flux due to clouds is thus 102 Wm^{-2} and is referred to as shortwave CRF. Additionally, because the effect was measured at the surface, it is referred to as surface shortwave CRF. This nomenclature will be referred to repeatedly in the discussion below.

Using satellite data from the Earth Radiation Budget Experiment, Ramanathan and Collins (1991; hereafter RC) provided a more quantitative analysis of the one-dimensional radiative feedbacks associated with regulating/limiting SST, the results of which led to their so-called "thermostat hypothesis". RC hypothesized that solar shielding due to convectively-generated, highly-reflective cirrus clouds was the primary limiting mechanism on SST growth, and therefore acted as a "thermostat" for the warmest tropical regions. Their hypothesis was based on three premises. First, observations and model calculations show that the downward longwave emission from the atmosphere to the surface begins to grow at a faster rate than the upward longwave emission from the surface in the range of SST greater than about 25^{o}C. This is primarily due to increased concentrations of, and enhanced radiation trapping effects from, atmospheric water vapor (Raval and Ramanathan, 1989, Stephens and Greenwald, 1991). This phenomena is often referred to as the "super greenhouse effect" and results in the longwave radiation component ceasing to function as a negative feedback on SST growth in the upper range of tropical SSTs (i.e., $\partial L/\partial T_S < 0$ for SST > ~25^{o} C). In the absence of other negative feedbacks, this behavior would result in a "runaway" greenhouse effect. Second, RC found that local changes in top-of-the-atmosphere (TOA) shortwave CRF occurring in association with tropical Pacific SST changes during the 1985-87 ENSO "cycle" indicated clouds could act as a negative feedback with a magnitude of about 22 $Wm^{-2}K^{-1}$. An illustration of how this feedback might come about in the process of El Niño warming is given in Figure 5. Third, RC argued that surface evaporation could not function as a limiting mechanism in the warmest oceans because evaporation adds moisture to the atmospheric column, enhancing the super greenhouse, and thus, diminishing the cooling effect from the net longwave flux. RC also noted that water vapor is actually imported into the warmest ocean regions and that evaporation tends to be low in these

[1] Collins et al. (1996) discuss in significant detail the radiative effects of convection and illustrate a relationship analogous to Fig. 3 using shortwave CRF instead of deep convection.

areas. Given these facts and inferences, RC concluded that SST-induced changes in highly reflective cirrus clouds associated with deep convection accounted for the negative feedback which limited SSTs to about 30-32^{o}C.

The three components of RC's hypothesis each tend to modify aspects of the simplified surface heat budget model proposed by Newell which was later modified by GB. The first observation regarding the behavior of the net longwave requires that the curve labeled L in Figure 3 begin to decrease, around an SST equal to about 25^{o}C. The second inference regarding the behavior of the CRF essentially requires that the insolation curve be sloped such that it decreases with increasing SST at a rate of about 22 $Wm^{-2}K^{-1}$. The last inference regarding the behavior of the evaporative heat flux suggests that it is sloped too steeply in the Newell model, and in fact brings to question the sign and strength of its feedback (i.e., $\partial E/\partial T_S$) at the higher SST values. A schematic depiction of the revised behavior of the surface energy budget analogous to Figure 3 is given in Figure 6. In accordance with the open question concerning evaporation, its value in this depiction is simply set to a constant consistent with its observed mean in the warmer tropical regions [e.g., Weare et al. (1980); see Zhang and McPhaden (1995) below]. Note that in the RC model, the braking mechanism on SST is now primarily due to decreasing insolation associated with increasing clouds as SST increases.

Since RC's "thermostat hypothesis" was put forward, great interest has surrounded the issue of SST regulation/limitation and a significant number of additional studies have been undertaken. These studies have applied innovative analyses and often offered alternative mechanisms and conclusions regarding the manner our warmest climates are regulated. One uncertainty that continues to be addressed through these studies involves the behavior and relative roles of surface flux feedbacks associated with increasing SSTs, mainly those associated with latent, shortwave and longwave fluxes. A significant part of this uncertainty arises because observational data for these surface fluxes are somewhat suspect either because they are based on sparsely sampled in-situ measurements (e.g., Weare et al., 1980) or because they are derived from satellite measurements. How to translate the latter of these into geophysical quantities at the surface poses a very challenging problem (e.g., Liu, 1988; Pinker et al., 1995; Cess et al., 1995). Even more uncertain and under debate are: 1) the manner the locally-high SST environment and associated surface fluxes couple to the large-scale atmospheric circulation, and 2) the role of ocean processes in regulating/limiting SST. These two questions have arisen to some extent due to the above studies' local framework of analysis where the atmosphere was viewed as a column in which the associated SST, insolation, cloudiness and other heat loss terms only interacted locally, and the interaction between adjacent columns as well as the interaction with oceanic processes was not explicitly taken into account. The continued uncertainty associated with these questions stems from the difficulty of analyzing local-to-remote atmospheric coupling as well as addressing coupled ocean-atmosphere processes. In the next section, a brief description is provided of some of the more recently advanced results and hypotheses regarding the role of the large-scale atmospheric circulation in regulating/limiting tropical SSTs. Section 5 follows with an analogous discussion concerning studies which have explicitly involved or invoked the role of the ocean. These discussions are not meant to be exhaustive[2] but are intended to emphasize the complex nature of this question, the progress that has been made, and the significant amount of work that remains.

4. Coupling to the Three-Dimensional Atmospheric Circulation

Within the context of this renewed interest in regulating/limiting mechanisms for tropical SSTs, one of the first to emphasize the role of the large-scale circulation was Wallace (1992). In his response to RC's "thermostat hypothesis", Wallace used a heuristic argument to suggest that cloudiness is not necessary to limit SSTs due to the efficient re-distribution of heat in the horizontal by the atmospheric circulation. If one assumes a homogenous atmosphere and ocean in equilibrium and superimposes large-scale "hot and cold SST patches" of equal magnitude into the system, Wallace argued that enhanced heat fluxes over the hot patches would remove heat from the ocean, producing an overlying unstable atmosphere. This will lead to enhanced deep convection and a rapid redistribution of heat through the resulting horizontal circulation. The atmosphere over the cold patches, on the other hand, would be hydrostatically stable with the result that the cold patches would be more resistant to change through perturbations by air-sea fluxes. This led Wallace to infer that the interactions between surface evaporative/sensible heat flux and the atmospheric circulation would lead to the observed skewness in the SST population distribution and alone be sufficient to limit SSTs in the tropics to within about 2°C of the most typical value (see Figure 2).

An observational indication of how remote influences from the large-scale atmospheric circulation can play a role within the context of high SSTs came from further studies of the relationship between deep convection and SST. With the advent of significantly more data than was used in the GB study, Waliser et al. (1993) resolved the high SST limit of the relationship between SST and deep convection for monthly and longer time scales (see Figure 4). They found that above about 29.5°C, the intensity/frequency of monthly-averaged deep convection decreases with increasing SST, and thus, the highest observed SSTs tend to be found in areas or at times of suppressed deep convection. Waliser and Graham (1993; hereafter WG) examined the data further and found that suppressed (enhanced) deep convection tends to be associated with positive (negative) rates of SST change on the order of 0.1°C month^{-1}. Based on these results, those in Figure 4, and the results from a simplified surface heat budget model[3], WG concluded that: 1) when deep convection is suppressed in warm-pool regions by remotely forced atmospheric subsidence, areas of very warm SST (> 30°C) can form, and 2) these high SSTs are unstable and lead to probable increases in local deep convection which cool the SST through enhanced cloudiness and turbulent heat exchange. The first of these implies that the coupling relationship changes from one that is dominated by the ocean forcing component for SSTs less than 29.5°C (e.g., increased SST lead to enhanced deep convection), to one that is dominated by the atmospheric forcing component for SSTs greater than 29.5°C (e.g., enhanced atmospheric subsidence leads to increased SSTs). WG also attempted to address whether the convective

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[2] Additional studies related to this topic include: Sarachik, 1978; Betts and Ridgway, 1981; Lindzen, 1990; Heymsfield and Miloshevich, 1991; Sui et al., 1991; Stephens and Slingo, 1992; Sun and Lindzen, 1993; Arking and Zisken, 1994; Inamdar and Ramanathan, 1994; Liu et al., 1994; Webster, 1996; Waliser, 1996a.

[3] The model was an extension of Newell's but with convection/cloudiness dependent on SST (e.g., Fig. 3), wind speed dependent on the convection, a more realistic form of the net longwave formulation (e.g., Fig. 5), and a more realistic parameterization of the air-sea temperature and specific humidity differences (e.g., T_s-T_a in Eq. 2).

perturbations to the surface shortwave or latent heat fluxes were larger, and concluded from limited data and the simplified heat budget model that in the local area of high SST, the cloud shielding effect on shortwave radiation appears to be more important (cf. Young et al., 1992).

Zhang and McPhaden (1995) examined the relationship between evaporation and SST in a manner analogous to that shown for convection in Figure 4. Their observational results, reproduced in Figure 7, show that evaporation tends to decrease with increasing SST, for SST above about 28ºC. They found that this decrease is driven by a decrease in wind speed (see Equation 1) which balances or even overcomes the increase in specific humidity difference due to the rise in Q_S via SST. Based on additional data analysis and model simulations, Zhang et al. (1995) hypothesize that this is related to the large-scale effects of convection through the following process. A region of large-scale high SST will be convectively unstable and induce upward motion, which by mass continuity brings about low-level convergence. In the center of this convergence, the wind speed is inherently low, yielding relatively low surface latent heat fluxes. In partial agreement with the last of RC's premises, these results suggest that evaporation is not, at least to the degree suggested by Newell (curve E in Figure 3), the limiting mechanism for a local area of high SST.

To further illustrate the role of the large-scale circulation as well as determine the relative contributions to the surface heat budget from the evaporation and surface shortwave perturbations, Waliser (1996b) used a multitude of data sets to describe the three-dimensional evolution of the atmosphere-ocean coupled system associated with the development of very high SST regions in the tropical western Pacific. Using satellite data, historical atmospheric and ocean assimilation products and in-situ data, Waliser composited ocean and atmosphere conditions for the months prior, during and after the development of ocean "hot spots" to examine the manner these high SST regions develop and then decay. These hot spots were defined as regions with monthly SST values greater than 29.75ºC (i.e. the upper tail of the distribution shown in Figure 4) and an area larger than 10^6 km^2. Results based on hot spots from the western Pacific warm pool (0-10ºS; 156ºE-176ºW) indicate strong influences from interannual, annual and 30-60 day time scales, with La Niña conditions appearing to inhibit formation, southern summer favoring formation, and the descending (ascending) phase of the Madden-Julian Oscillation (MJO) favoring formation (decay). With respect to the surface heat budget, the composites showed that during the hot spot evolution, the convective perturbations to the surface shortwave exceed those for evaporation by at least a factor of two. This observational analysis elicits a comparison between the composite "hot spots" and the hypothesized "hot patches" of SST invoked in the arguments of Wallace (1992) described above. Following the argument of Wallace, hot patches will be subject to enhanced fluxes of latent heat. This enhanced latent heat flux cools the SST and increases the frequency of convection over the hot patches. The convection then distributes the heat over the entire tropical troposphere. The composites show that while convection does increase over the hot spot and the temperature of the troposphere is increased as a result, the increases to the latent heat flux are quite modest, with the perturbations to the surface shortwave appearing to be more important to the SST cooling. Thus, with the exception of the relative importance between shortwave CRF and evaporation, the behavior of the hypothesized "hot patches" is similar to the observed "hot spots".

Using a simplified model of the tropical atmospheric boundary layer, Hartmann and Michelsen (1993; hereafter HM) provided an alternative mechanism for the regulation of tropical SSTs. Their model was based on the coupling of three equations: 1) a surface energy budget equation which included a specified ocean transport; 2) a boundary-layer moisture budget equation used for the calculation of the evaporation; and 3) a boundary-layer, zonal-wind anomaly equation based on the SST-gradient scheme of Lindzen and Nigam (1987). A schematic depiction of their model framework is given in Figure 8. The warm region to the left represents the western Pacific warm-pool and the cool region to the right represents the eastern Pacific cold tongue. The resulting gradient in SST between the east and west drives a westward zonal wind anomaly which advects air from the east into the western region. From Equation 1 above, it is evident that this cool, dry, boundary-layer air acts to enhance the evaporation and reduce the local SST along with the zonal SST gradient. Furthermore, the enhanced zonal wind has the effect of increasing evaporation over the whole domain leading to a decrease in the domain-averaged SST as well. While HM acknowledge the important role that shortwave CRF may have over the area of highest SSTs (those in the west), they show that inclusion of a linear negative shortwave feedback in their model, similar to that shown in Figure 6, would actually increase the domain-averaged SSTs. In partial support for this last point, HM used Earth Radiation Budget Experiment data to show that while shortwave CRF increased over the anomalously warm SST region associated with the El Niño analyzed by RC, the change averaged over the entire Pacific Ocean was nearly zero. This led HM to suggest that it is the coupling between the evaporation, large-scale circulation, and SST gradients that regulate the tropical SST (i.e., determine the SST mode or the tropical average SST). Fu et al. (1992) performed a similar analysis to compare the CRF feedbacks between the local versus remote environments and found analogous results. In addition, they showed that when evaporation is averaged over the entire tropical Pacific for the same ENSO period analyzed by RC, it provides a negative feedback with a magnitude of about 20 $Wm^{-1}K^{-1}$, even though it is much smaller or even acts as a positive feedback in the local area of highest SST anomaly (e.g., Liu and Gautier, 1990; Zhang and McPhaden, 1995).

Recently, Zhang et al. (1996, hereafter ZET; cf. Chou, 1994) extended the portion of HM's analysis concerning ENSO-related changes to domain-averaged CRF. ZET computed the domain-average TOA longwave, shortwave and net CRF anomalies from ERBE data along with the corresponding domain-averaged SST anomaly. ZET then computed the linear regressions between the CRF anomalies and the SST anomalies which could be interpreted as a measure of the cloud radiative feedbacks associated with a given SST perturbation. The regressions were computed for domain sizes increasing from a relatively small region in the western Pacific to the entire global tropics from 30ºN to 30ºS. Their results, reproduced in Figure 9, show that the ratio with the shortwave CRF indicates a net cooling for the smaller region, then monotonically increases to the point where it indicates a net warming for the global tropical region. Consistent with HM, the zero value for the shortwave-SST regression corresponds with the tropical Pacific domain. The regression for the longwave CRF has just the opposite behavior but is slightly larger at both extremes. The regression for net cloud radiative forcing with SST is positive (negative) for smaller (larger) domains, indicating induced warming (cooling). ZET report that this behavior is largely the result of changes in cloudy-sky longwave emission coupled to the large-scale circulation.

At this point some clarification is in order to account for the differences in the shortwave CRF feedbacks computed by RC and ZET. As indicated above, the regressions computed by ZET result from averaging CRF and SST anomalies (about the annual cycle) over a given region for a given month and then plotting these points on a scatter diagram, one point for each month of ERBE data. From these points a linear regression is computed. Note that as the domain in ZET's analysis shrinks, the CRF regression values increase in magnitude. The values for the smallest domain (i.e., a single grid point) will depend on the climatological location, i.e., the sensitivity of that location's CRF to changes in SST. RC computed shortwave CRF feedback by taking anomaly values from each grid point within a region (10ºN-10ºS; 90ºE-140ºW, similar to one of the smaller ZET domains), computing the differences between El Niño and La Niña years, plotting each of these points on a scatter diagram, and then computing a linear regression[4]. This method is somewhat similar to the ZET method but for a grid-point size domain and thus leads to relatively large feedback values (22 $Wm^{-2}K^{-1}$), especially since the "grid-point domains" came from a climatological location that is particularly sensitive to SST changes (i.e., central tropical Pacific).

The above observational analyses have obvious connections to Pierrehumbert's (1995) "radiator fin" hypothesis for the regulation of tropical SSTs. This hypothesis stems from the development and consideration of an idealized radiative-dynamic model of the tropical circulation. While the model is highly simplified and requires the imposition of several important assumptions, it does take into account the physics of the atmospheric column, including the coupling of an ascending cloudy region and a descending cloud-free region (e.g., Hadley/Walker circulation). The model behavior indicates that evaporation on its own cannot provide an effective surface temperature regulating mechanism, and in addition, that clouds cannot serve as regulators either unless there are substantial departures between the observed cancellation between TOA cloud greenhouse and albedo effects[5]. In fact, the author finds that the main determinant of tropical temperatures stems from the efficiency through which the troposphere can radiate its heat to space. This is dependent on the size and dryness of the subsidence regions in the tropics. In these regions, referred to as "radiator fins", the greenhouse trapping due to water vapor and clouds is weakest, and thus, the emission of longwave energy from the ocean-atmosphere column is greatest. In the absence of these radiator fins the tropical temperature would increase dramatically and be subject to a runaway greenhouse effect.

In another effort to illustrate the important role the large-scale circulation has on cloud cover, and thus the negative feedback it implicitly has on SST, Lau *et al*. (1994; hereafter LET) posed a hypothetical numerical experiment using a highly sophisticated regional cloud model. LET performed three experiments to test the relative effects on local shortwave CRF due to: 1) local SST changes, and 2) remote atmospheric circulation changes which induce localized atmospheric vertical motion. In the control experiment, the boundary conditions consisted of an SST of 28ºC and an imposed upward vertical velocity to account for remote circulation effects. In the first perturbation experiment, LET raised the SST to 30ºC and left the imposed vertical velocity the same as the control case. In the second perturbation experiment, LET removed the imposed vertical velocity, thus shutting off the remote influence, and left the SST the same as the control case. Comparing the shortwave CRF from the two perturbation cases to the control case, LET found that the shortwave CRF was significantly more sensitive to the imposed changes in the "remote" circulation than the imposed change in local SST of 2ºC. While it is not obvious that

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[4] In addition, RC also multiplied each point's x-axis (ΔSST) and y-axis (ΔCRF) values by their associated ΔSST. Since the accuracy of a given SST observation is thought to be about 0.5ºK, this reduced the impact of noise on the regression stemming from points with very small ΔSSTs.

[5] Even though a cloud anomaly cools the surface by reducing the surface shortwave flux and heats the troposphere by trapping a portion of the longwave flux emitted by the surface, observations at the top of the atmosphere show that the reduction in outgoing longwave flux by the cloud is equal to the reflection of shortwave flux by the cloud (see Kiehl, 1994). Thus, the ocean-atmosphere column undergoes no net change in total energy flux.

these two perturbation experiments are commensurate tests of the two mechanisms being examined, the results do provide support for LET's assertion that a better understanding of cloud feedbacks in regards to their role in regulating/limiting SST will only come by considering both local and remote influences.

5. Incorporating Feedbacks from the Ocean Circulation

In the studies discussed in the section above, ocean processes are either ignored completely or only passively taken into account via a simplified surface energy budget with a fixed ocean heat transport. In a few cases, studies addressing the role of shortwave cloud feedbacks and the role of the atmospheric circulation have been performed in a context which explicitly accounts for a more accurate depiction of ocean processes. In one such study, Waliser et al. (1994) examined the role of cloud shortwave feedbacks in the regulation of ENSO. The motivation for the study was based on the observation that during a La Niña period, the SST in the central Pacific is relatively cool which results in anomalously weak convection. This allows more shortwave heat flux to penetrate the surface. The opposite conditions exist during the warming of an El Niño period (e.g., Figure 5). Considering this cloud shortwave effect alone, each of these conditions would tend to move the SST toward a state in between the two extremes. Using coupled numerical ocean simulations, Waliser et al. found that a simplified, empirical parameterization for this SST-induced shortwave feedback acted to limit the geographical extent and magnitude of the temperature extremes associated with the simulated ENSO, indicating the possible importance of such a feedback in limiting/regulating tropical SSTs. However, because the ENSO simulation was qualitatively correct without this feedback, the authors concluded that such a negative feedback was not critical to the first-order regulation of ENSO. The one caveat to this conclusion given by the authors involved the parameterization of the surface evaporation, which did not include active evaporation-wind feedback processes (thus the U in Equation 1 was fixed in time for these simulations). Observations show that low mean wind speeds are typically associated with high SSTs. These low wind speeds limit the effectiveness of the evaporation as a cooling mechanism (e.g., Zhang et al., 1995), particularly in the case of ENSO (e.g., Liu and Gautier, 1990). If this latter process was modeled more realistically, the shortwave cloud feedback may become more significant with respect to SST regulation of the simulated ENSO.

To test the sensitivity of the so-called "cloud albedo feedback" mechanism[6] for regulating and/or limiting tropical SSTs, Meehl and Washington (1995; hereafter MW) employed a coupled global circulation model. The coupled model contained a 3-layer thermodynamic sea-ice model, a 20-layer ocean model and a 9-layer atmospheric model. The latter computed its own cloud amounts via a penetrative convective scheme. Rather than use fixed optical properties for clouds, a cloud albedo scheme was incorporated to account for the brightening of clouds as convection increases. In this parameterization, the albedos were linearly increased when SST rose above a specified threshold temperature. Given the inclusion of this parameterization, the model contained two competing feedback loops, one that diminishes climate sensitivity and another that increases climate sensitivity. The former arises from the cloud albedo feedback, where any perturbation, such as an increase in convection, will lead to more moisture in the troposphere, increased clouds, slightly elevated albedo (in the region where SST is above the

[6] The idea that clouds become more reflective and occur more frequently as SST increases (e.g., Figs. 4 and 5).

threshold), decreased surface shortwave flux, decreased SST, and thus a tendency to reduce the initial convection perturbation. The latter arises from the super-greenhouse effect, where a similar perturbation will lead to more moisture in the troposphere, increased clouds, increased downward longwave emission, increased SST, and thus, a tendency to enhance the initial perturbation. To test the sensitivity of the global climate to an increase in the effectiveness of the cloud albedo feedback, MW integrated the model with the threshold set at 303K for 38 years producing a stable and realistic climate simulation, and then followed this by reducing the threshold to 300K beginning at year 21 of the previous simulation. After a five-year integration, it appeared that, at least in this simulation, the cloud albedo effect dominated. The reduced threshold value produced a considerably cooler climate (~ 1.7°C decrease in globally averaged surface air temperatures) with accompanying widespread dynamical changes. MW stressed that the above experiments demonstrate how sensitive the (simulated) global climate can be to a regional change in cloud radiative forcing.

Just as much of the later work on the atmospheric side of the SST regulation problem has invoked the role of the large-scale circulation, the most recent and intriguing research in this area has been in considering the large-scale circulation of the ocean. Two recent efforts [Clement et al. (1996) and Sun and Liu (1996); cf. Dykstra and Neelin (1995)] have examined the effects of coupling equatorial Pacific ocean dynamics to essentially the simplified atmospheric model illustrated in Figure 8 which was discussed in the context of the HM study. In the latter, the effect of the east-west SST gradient was examined in terms of its ability to induce easterly winds and enhance the evaporation rates. In these two more recent studies, the effect of the SST gradient is examined in terms of the induced easterly winds' effect on ocean vertical upwelling and horizontal transport of cool water into the region(s) of warmer SST.

Clement et al. (1996; hereafter CET) used a simple coupled model of the tropical Pacific basin with an imposed seasonal climatology (Zebiak and Cane, 1987) to illustrate how ocean dynamics influence the response of the ocean mixed-layer (in their case 50m) when a uniform surface heating/cooling perturbation is imposed. For example, CET show that when the uniform heat flux of 40 Wm^{-2} is applied, the SST increases by less than 0.5°C averaged over the basin. However, in the case where ocean dynamics are ignored, the same forcing would produce about a 2°C increase in SST for the same mixed-layer depth. The uniform heating, in conjunction with the imposed climatology (i.e., east-west SST gradient), induces larger SST changes in the western and off-equatorial Pacific regions than the eastern equatorial Pacific where vertical transport of cool water up from the thermocline via upwelling can partly balance the additional surface heat flux. The resulting spatial structure of this SST response increases the east-west SST gradient. This reinforces the easterly wind stress, which increases the vertical upwelling in the east and the zonal transport of cool water to the west. Similarly, when a surface cooling perturbation is applied, the zonal transport of cool water weakens and the system cools significantly less than would be expected in the absence of the ocean dynamics. Therefore, the inclusion of this ocean dynamical feedback within the coupled ocean-atmosphere system significantly mediates the impact of an imposed heating/cooling, and thus, would appear to be a critical component in the regulation of tropical Pacific SSTs.

Sun and Liu (1996; hereafter SL) used a highly idealized box model to point out that the inclusion of the coupled dynamics discussed above lowers the climatological temperature of the warm region of their model ocean. A schematic of SL's box model is shown in Figure 10. In the absence of ocean currents, the temperatures in both surface boxes would equilibrate to a temperature determined by simple radiative-convective interactions with the overlying atmosphere (thermodynamic coupling). If for example, the temperature of the box in the west becomes slightly larger than that in the east, an SST gradient and associated low-level easterly flow are set up (Figure 8; dynamic coupling) driving a westward zonal current. In this case, the temperature of the ocean surface box in the east (T2) is additionally influenced by the outflow of water to the west and the inflow of water from beneath the thermocline. The size of this pool of water, and thus its heat capacity, is so large that its temperature can effectively be set to a constant ($T_c \sim 18$ºC). In a similar way, the temperature of the ocean surface box in the west (T1) is additionally influenced from the outflow to the region below the thermocline and the inflow of water at temperature T2 from the east. The results of their model show that as the dynamic coupling is increased relative to the thermodynamic coupling the model quickly enters into a state where both T1 and T2 are significantly lower than T_e. SL show that these diminished values are relatively insensitive to the specification of T_e, and even in the case of T_e estimated without cloud feedbacks ($T_e \sim 51$ºC), T1 and T2 equilibrate to about 30ºC and 25ºC, respectively. SL also show that when the same surface boundary conditions are applied to a general circulation model of an idealized Pacific Ocean basin, the behavior is nearly identical to that of the box model.

While the focus of the above investigations is primarily on coupled interactions between the ocean and atmosphere, the following results demonstrate how in some tropical areas it is necessary to consider interactions between the ocean, atmosphere and land components of the Earth's climate system to understand the manner in which upper limits are placed on SST. In the Indian Ocean, SST often achieves even larger values than what is observed in the Pacific. These large SST values are tightly coupled with the annual cycle, and primarily occur in the spring before the onset of the Indian summer monsoon (Figure 1). Figure 11 shows results from an analysis of ocean "hot spots", analogous to that discussed above by Waliser (1996b), except in this case, the compositing was done for very high SST values (SST > 30ºC) from the equatorial and northwest Indian Ocean. The four panels show the composites for total and anomalous SST, anomalous deep convection, and anomalous ocean wind stress. The top (bottom) map in each panel shows the composite field for the months during (after) the hot spots occurred. In all cases, the months the hot spots occurred were either April or May (compare SST panel with Figure 1b and note the elevated SST). Associated with the cooling in the month following the hot spot, is a strong signature of equatorial upwelling. This upwelling appears to be driven by an anomalous northeastward ocean wind stress which is coupled to the developing convection anomaly over the northwest Indian Ocean associated with the developing summer monsoon. This dynamical feature leads to enhanced advection and eddy fluxes of the sub-thermocline water into the surface layers where it acts to cool the SST. These results illustrate another example of how ocean dynamics, in this case coastal upwelling, play an important role in the regulation of high SSTs. In addition, it shows that in some instances it is necessary to consider the complexities of coupled air-sea-land interactions, such as the monsoon, in understanding the climate controls on high SSTs.

A recent study by Anderson et al. (1996; hereafter AET) has pointed out another feedback within the ocean mixed-layer which also might influence the maximum SSTs observed. Using a one-dimensional ocean mixed-layer model, AET point out that the mixed-layer contains a climate stabilizing feature itself. Typically, enhanced surface mixing of the ocean gives rise to surface cooling and a higher resistance to warming due to the deepening mixed-layer depth, while reduced mixing gives rise to just the opposite conditions. However, as the mixed-layer depth decreases in the latter case, more shortwave energy is actually deposited below the mixed-layer (Siegel et al., 1995). Since the heat loss terms (Equations 1-3) act to cool the surface of the mixed-layer, this gives rise to a hydrostatic instability which enhances the mixing and thus cools and stabilizes the system. As the mixed-layer grows, more shortwave energy is deposited in the mixed layer itself. With no change in the way the heat loss terms act, the mixed layer tends to shallow and warm. This albeit simple feedback provides a stabilizing effect which is not often treated even in sophisticated modeling scenarios due to the shortcomings still present in representing ocean mixed-layer processes.

6. Summary

This review describes many of the efforts taken to understand the mechanisms which control ocean surface temperatures in the Earth's warmest climates. These efforts have migrated from simple energy budget models to considerations of the radiative-convective nature of the atmospheric column, and have since examined the influences from the large-scale atmospheric circulation and the feedbacks from ocean dynamics. Further, they have provided a number of useful paradigms for considering the maintenance of tropical SSTs, both for the present-day climate as well as for one altered due to increasing greenhouse gas concentrations. The review strives to highlight the distinction between hypotheses and mechanisms which place upper bounds on tropical SST from those which act to regulate the tropical average SST. Such a distinction facilitates the assessment of their individual validity and overall coherence, and in addition, is useful when considering the local versus remote feedbacks associated with a given SST maximum and/or anomaly. Examination of an idealized scenario helps to illustrate this point and tie together many of the studies discussed above.

Consider the atmospheric response to the imposition of a large-scale positive temperature anomaly in an otherwise horizontally uniform ocean mixed-layer. In this case, the location of the anomaly will represent a localized SST maximum. In this region, the system will initially try and equilibrate through enhanced latent, sensible and net longwave surface fluxes. These enhanced fluxes will initiate a low-level atmospheric convergence over the region of maximum SST (Lindzen and Nigam, 1987; Zebiak, 1986). Keep in mind however, that the amount of heat and moisture removed from the ocean in order to initiate this circulation results in a negligible change in the ocean mixed-layer temperature due to the drastic difference in the heat capacities of water and air. Associated with the low-level convergence are: 1) strong upward vertical motion over the region of warmest SST with compensating, weak downward motion elsewhere, 2) a reduction in the mean value of the surface wind speed over the region of warmest SST relative to the surrounding areas (since the wind direction must change sign across the SST anomaly), and 3) the transport of low-level moisture into the region of high SST.

By examining the impact of these changes on the surface heat budget, it is evident from the studies described above that the first and third of these act to decrease (increase) the incoming surface shortwave (longwave) flux over the high SST area relative to its surroundings (e.g., RC), while the second and third have a similar effect on the surface evaporative flux (e.g., Zhang and McPhaden, 1995). Therefore, in the local area of high SST, the only surface flux feedback which is negative, and thus helps to directly remove the local SST perturbation, is the shortwave CRF. However, because the SST anomaly has raised the basin-average temperature as well as produced a significant non-local response (e.g., Gill, 1980; Wallace, 1992; HM), it is critical that we consider the influence of the basin-wide feedbacks. In this regard, the evidence suggests a response that is almost the opposite to that for the locally-high SST area. In the surrounding area, the induced atmospheric circulation (i.e., subsidence) acts to increase (decrease) the incoming surface shortwave (longwave) flux (e.g., HM, ZET, Pierrehumbert, 1995), and increase the surface evaporative flux (e.g., Fu et al., 1992; Zhang et al., 1995). Thus, any basin-scale cooling that results comes about via changes in evaporation and net longwave fluxes more so than from changes in shortwave CRF. This idealized scenario emphasizes the interlocking and complimentary nature of many of the mechanisms and hypotheses that have been advanced to date regarding tropical SST controls, and although illustrative, it still leaves out the role of the ocean.

Incorporating ocean feedbacks into the above simplification is difficult. In an uncoupled framework, ocean dynamics will distribute the SST anomaly throughout the domain, with a time scale somewhat dependent on the proximity of the initial anomaly to the equator (see Philander, 1991). This dynamical response in conjunction with some form of simplistic dissipation will lead to the removal of the anomaly over fairly broad time and space scales. In the ocean-atmosphere coupled case, the equilibrium of this idealized scenario is nearly synonymous with the physics of ENSO, at least in the case of the anomaly occurring in the tropical Pacific. Regulation of SST within the context of this problem has been studied extensively and involves ocean wave dynamics, vertical upwelling and mixing processes, surface heat exchange, and coupled SST-wind feedbacks (see review by Neelin et al., 1994). While there is no clear connection between the mechanisms which regulate ENSO-related SST variability and those which limit tropical SSTs, ENSO variability has often been invoked to test many of the mechanisms described in this review (e.g., RC, HM, Waliser et al., 1994). Experimentally useful scenarios that lie outside the ENSO case are rare in the observed data, especially for the ocean where data is significantly more sparse. The idealized scenarios given by CET and SL are certainly applicable to the case of greenhouse warming where a weak and relatively uniform increase in downwelling longwave radiation might be expected (i.e., increase in T_e in the case of SL). As they each show, the response of the ocean can be quite important in regulating basin-averaged SSTs as well as placing a limit on the maximum temperatures observed.

In summary, the studies discussed in this review have yielded a number of important results and useful hypotheses regarding the control of tropical SSTs, generally through the analysis of limited subsets of observed data or the application of idealized modeling scenarios. Further progress will be achieved by focusing on the connections between the various mechanisms and their different regimes of applicability. Such an approach was hinted at in the idealized scenario above but needs to be expanded through analysis of more comprehensive observational data sets (e.g., Waliser, 1996b) and more focused use of fully-coupled, ocean-atmosphere models (e.g., MW, Schneider et al., 1996). Finally, it is worth emphasizing that much of the progress obtained thus far has relied heavily on the resource, data and programmatic support provided by many of the recent large-scale field and operational programs, such as the internationally sponsored Tropical Ocean Global Atmosphere (TOGA) program and TOGA Coupled Ocean Atmosphere Response Experiment (COARE), the United States National Science Foundation's and Department of Energy's Central Equatorial Pacific Experiment (CEPEX), and the United States National Oceanographic and Atmospheric Administration's TOGA Tropical Atmosphere Ocean (TAO) buoy array program. Further progress hinges on continued support in these areas, especially in the way of high-quality, long-record climate monitoring programs.

Acknowledgments: Support for this work was provided by the Climate Dynamics Program in the Atmospheric Sciences Division of the United States National Science Foundation under grant ATM-9420833 and the Climate and Global Change Research Program of the United States National Oceanic and Atmospheric Administration under grant NA56GP0236. The author would like to thank De-Zheng Sun, Guang Jun Zhang and Minghua Zhang for supplying reproductions of their figures and Wufeng Zhou for help in other figure production.

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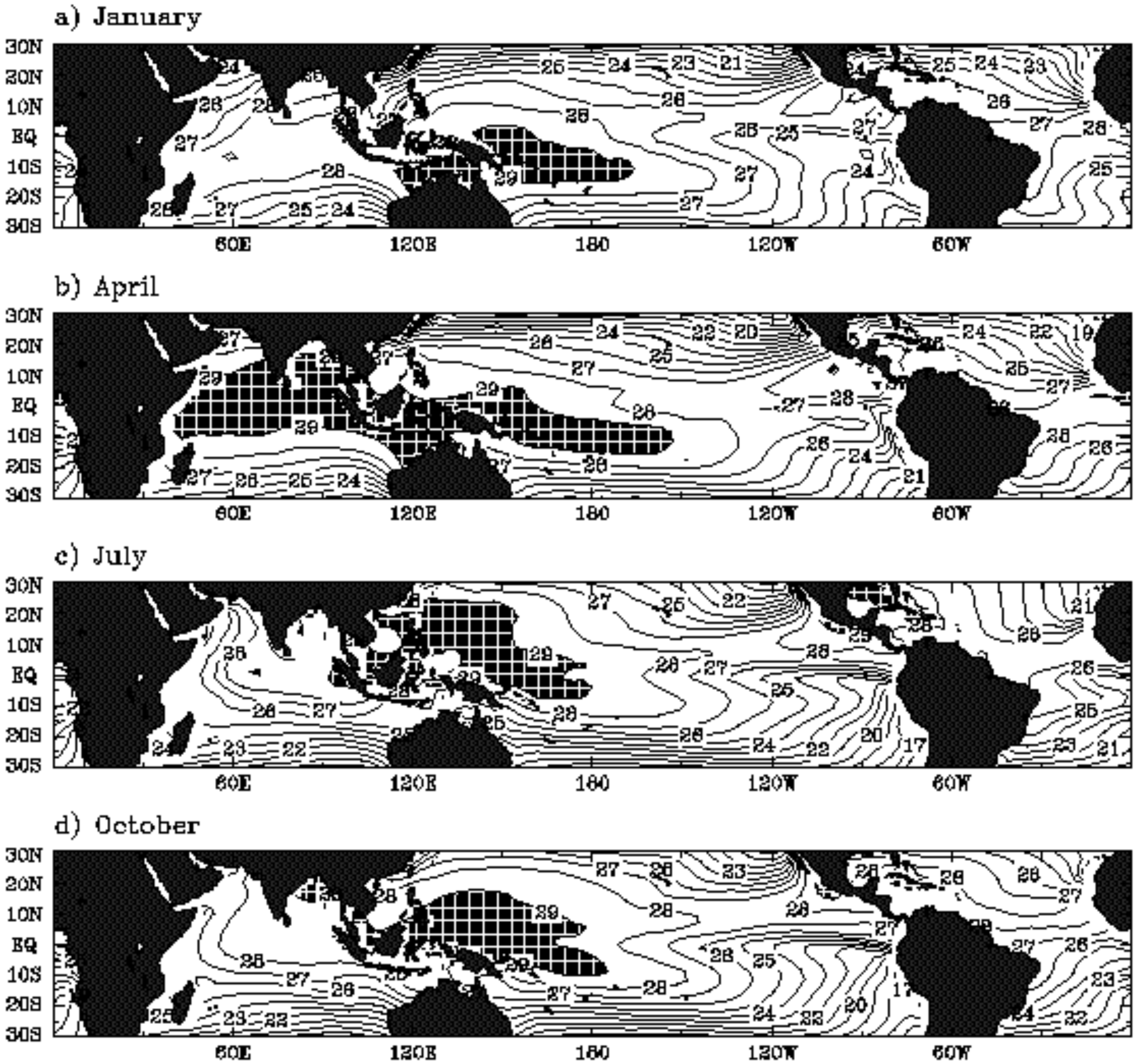

Figure 1: Monthly climatological sea surface temperature (SST) for (a) January, (b) April, (c) July, and (d) October. Shaded denotes regions where the SST is higher than 29ºC.

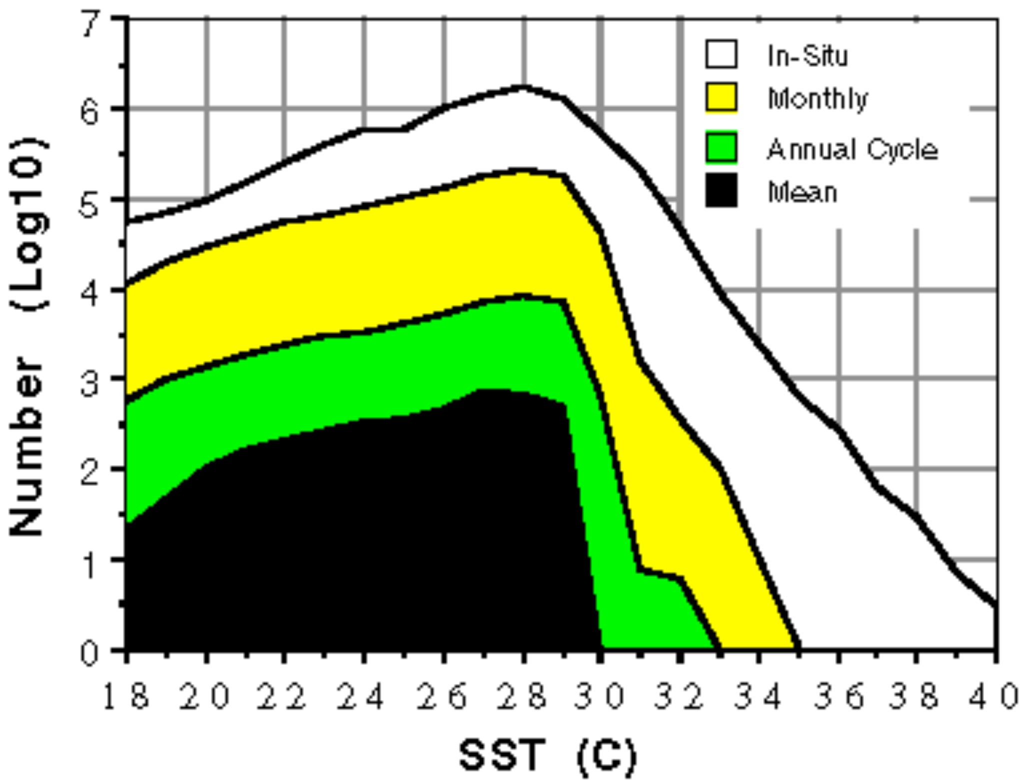

Figure 2: SST population distributions based on data from 25ºN to 25ºS. Distributions for the long-term mean (black), twelve-month annual cycle (dark gray), and monthly (light gray) data are computed from monthly data from 1970-1993 (Reynolds and Smith, 1994). Distributions for the in-situ (white) data are computed from Compressed Marine Reports (CMRs) from 1954-1979 (COADS, 1985).

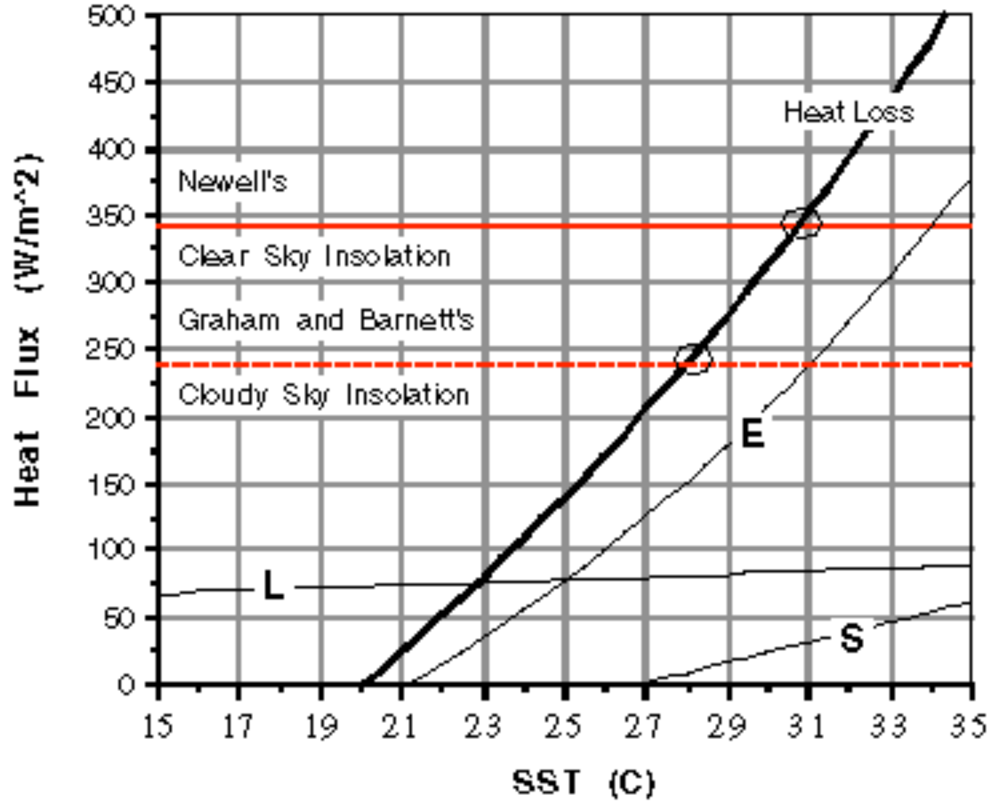

Figure 3: Simplified heat budget model of Newell (1979). Curves labeled E, L and S represent surface latent, net longwave and sensible heat fluxes, respectively; thick solid curve is their sum. Solid flat line is Newell's clear-sky insolation value. Dashed line is a reduced value of insolation accounting for the presence of clouds (Graham and Barnett, 1987). Circles denote SST where heat input and loss terms balance.

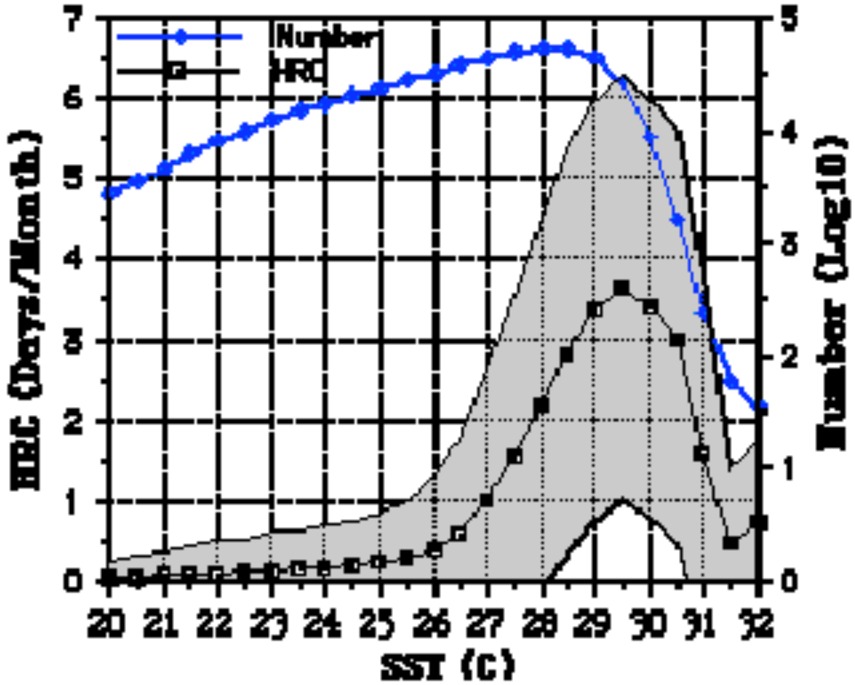

Figure 4: Mean HRC values for 0.5°C SST bins, computed from monthly values between 1975-87 for the global tropical oceans (25°N to 25°S). The number of values (line with circles) used in the computation of the means is specified by the right vertical axes. The standard deviation of the means are delineated by the shading. Reproduced from Waliser and Graham (1993).

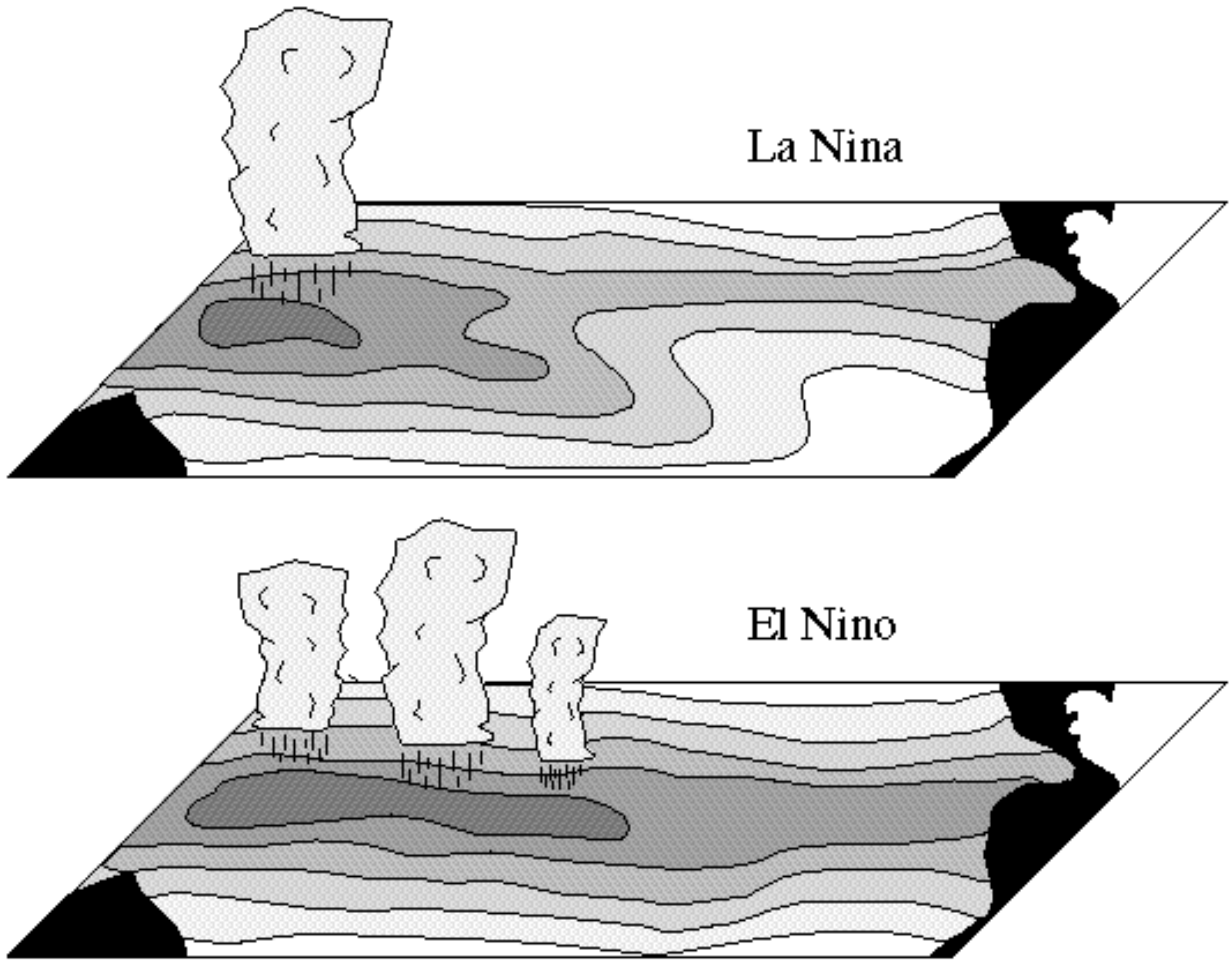

Figure 5: Schematic depiction of the Pacific Ocean SST and deep convection distributions for the two extreme phases of the El Niño Southern Oscillation. Top panel illustrates La Niña state (e.g., 1985), bottom panel illustrates El Niño state (e.g., 1987).

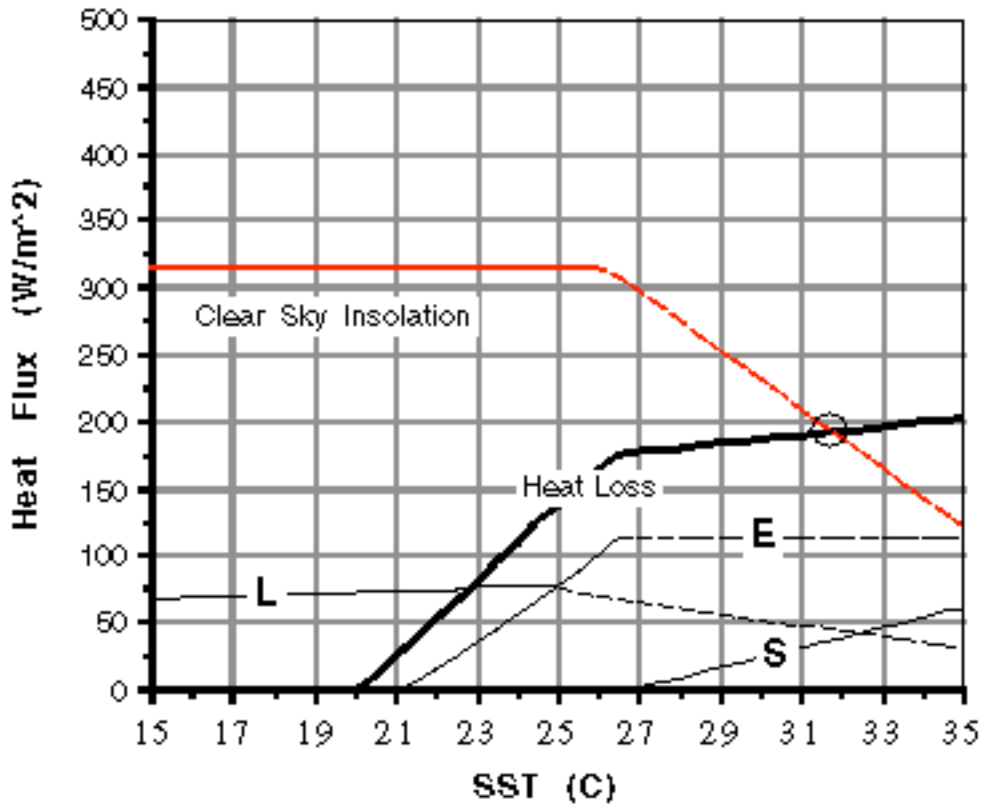

Figure 6: Same as Figure 3, except modified to illustrate the main elements to the "thermostat hypothesis"; see Section 3 for details. Modifications to the net longwave, evaporation and clear-sky shortwave are shown in dashed lines and reflect more recent observations and analysis from Kiehl and Briegleb (1992), Zhang and McPhaden (1995) and Waliser et al. (1996), respectively.

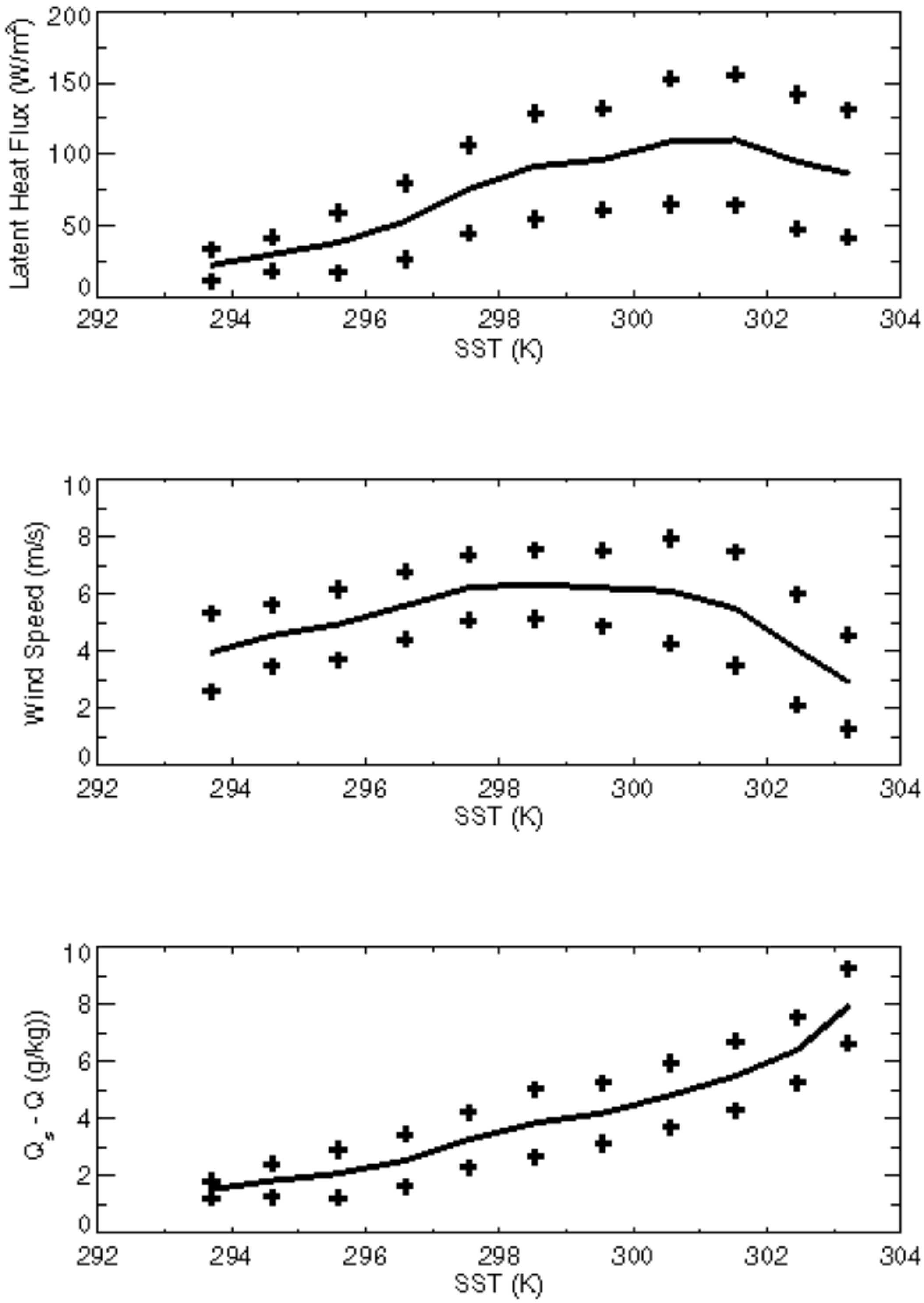

Figure 7: Surface latent heat flux, wind speed and humidity difference [$Q_s(T_s) - r\ Q_s(T_a)$] as functions of SST; see Equation 1. Mean values (solid line) and standard deviations (symbols) are plotted for 1ºC SST bins. Reproduced from Zhang and McPhaden (1995).

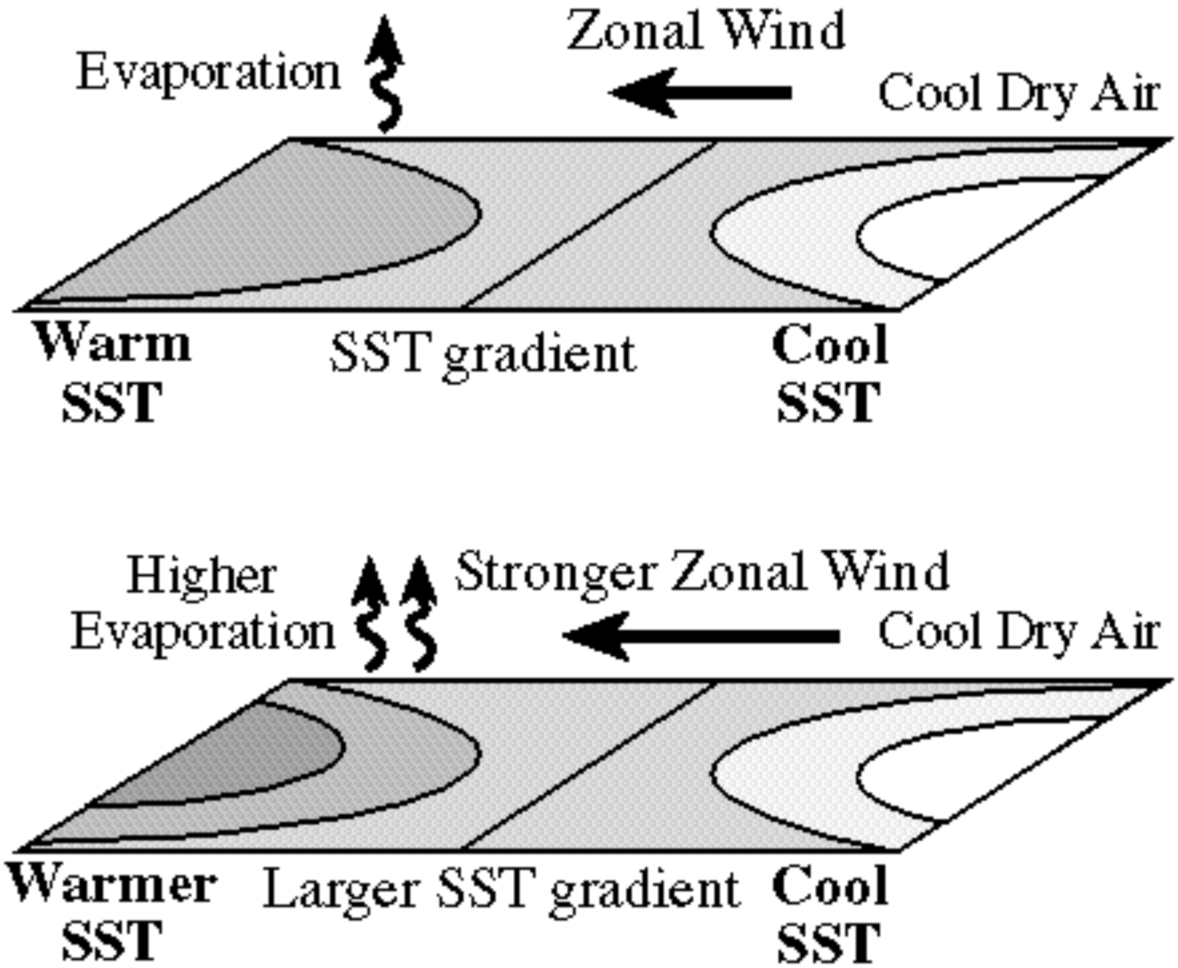

Figure 8: Schematic depiction of the main elements of the Hartmann and Michelsen (1993) mechanism for regulating tropical SSTs. See Section 4 for details.

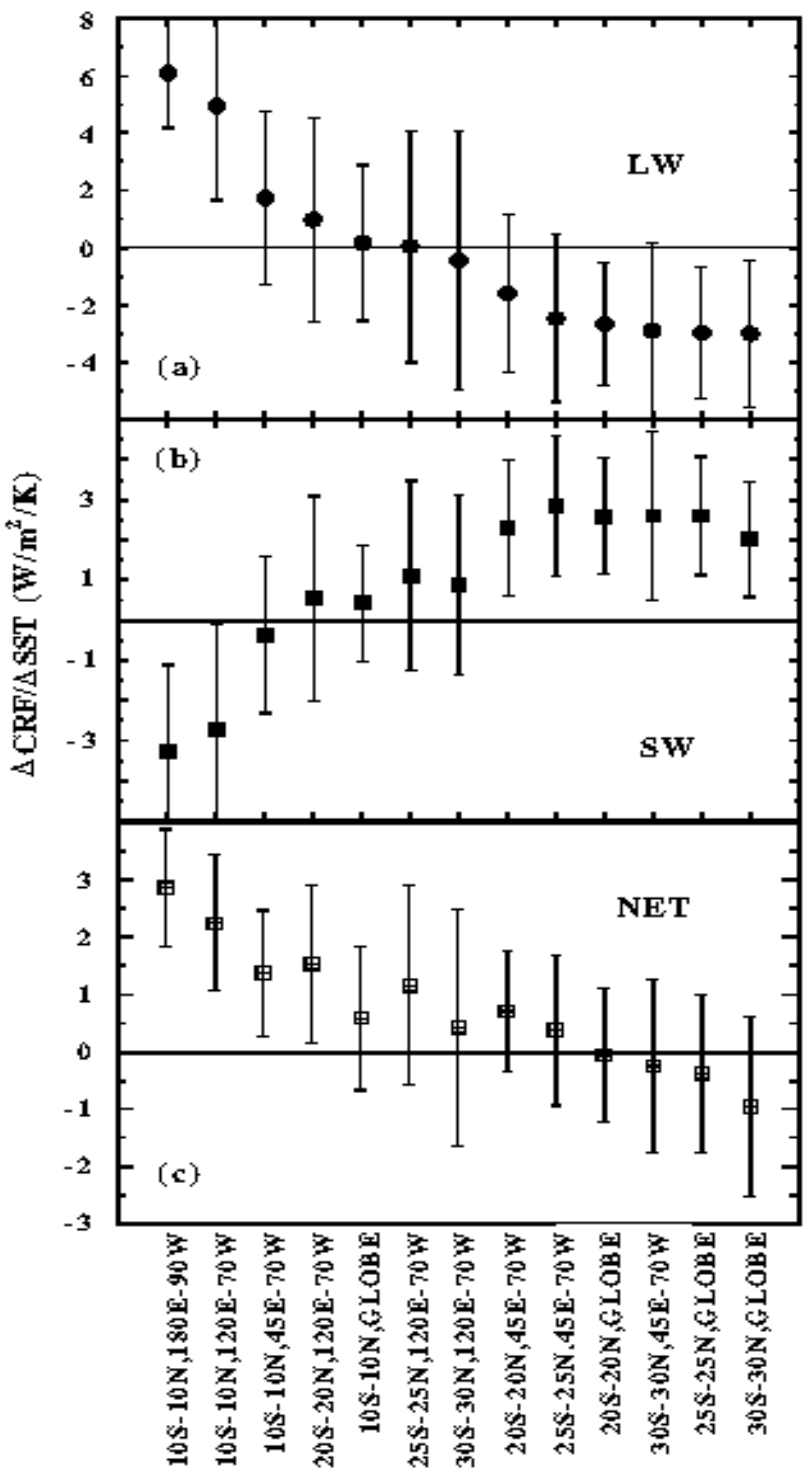

Figure 9: Regression ratios of the longwave (a), shortwave (b) and net (c) cloud radiative forcing (CRF) anomalies against corresponding SST anomaly. Horizontal axis denotes the domain the anomalies were averaged over before computing ratio. Error bars denote 95% confidence interval in (a) and (b), 80% in (c). Reproduced from Zhang et al. (1996), see Section 4 for more details.

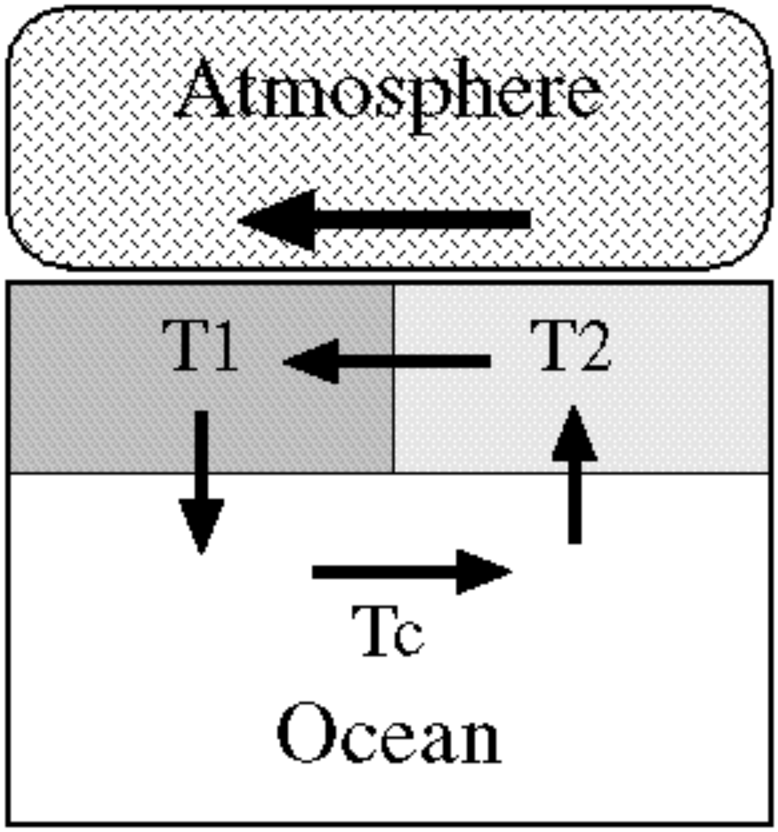

Figure 10: Schematic of three-box ocean model of Sun and Liu (1996). See Section 5 for details.

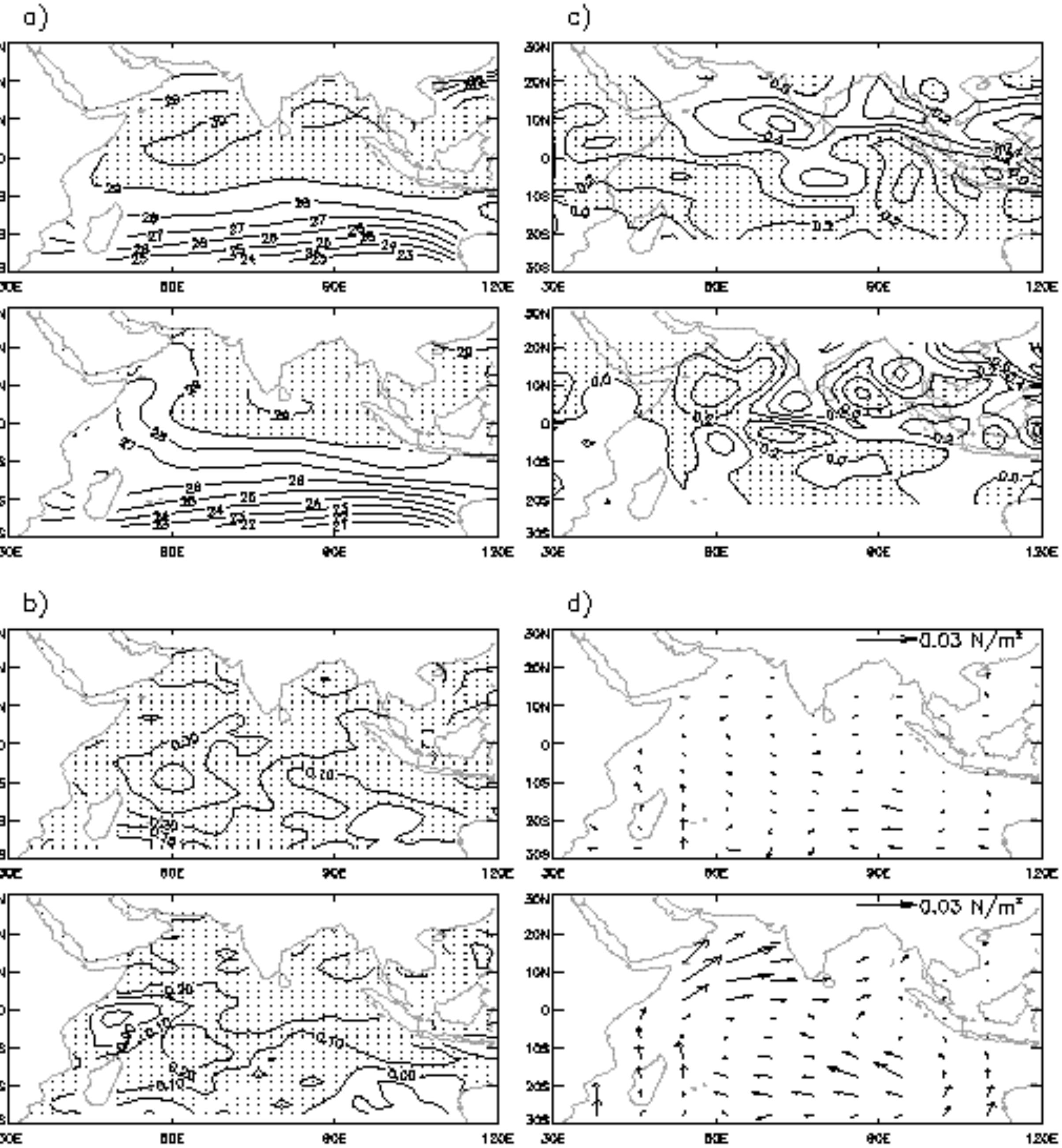

Figure 11: Composite analysis of SST (^{O}C; a), SST anomaly (^{O}C; b), Highly Reflective Clouds (days month^{-1}; c), and surface wind stress (Nm^{-2}; d). Top (bottom) panels are composite conditions during (after) very high SST conditions.